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Relative roles of stream flow and sedimentary conditions in controlling hyporheic exchange

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Abstract

Hyporheic exchange is often controlled by subsurface advection driven by the interaction of the stream with sedimentary pore water. The nature and magnitude of the induced exchange flow is dependent on the characteristics of both the stream flow and the sediment bed. Fundamental hydrodynamic theory can be applied to determine general relationships between stream characteristics, sediment characteristics, and hyporheic exchange rates. When the stream bed is fine enough to allow application of Darcy's Law, as with sand beds, the induced advective exchange can be calculated from fundamental hydrodynamic principles. Comparison with a wide range of experimental results demonstrates the predictive capability of this theory. Coarser sediments such as gravels are more complex because they admit turbulent interactions between the stream and subsurface flows, which can produce considerable exchange even when the bed surface is flat and no flows are induced by the bed topography. Even for this case, however, scaling arguments can still be used to determine how exchange rates vary with stream and sedimentary conditions. Evaluation of laboratory flume experiments for a wide range of stream conditions, bed sediment types including sand and gravel, and bed geometries demonstrates that exchange scales with the permeability of the bed sediments and the square of the stream velocity. These relationships occur due to fundamental hydrodynamic processes, and were observed to hold over almost five orders of magnitude of exchange flux. Such scaling relationships are very useful in practice because they can be used to extend observed hyporheic exchange rates to different flow conditions and to uniquely identify the role of sedimentary conditions in controlling exchange flux.

Introduction

As a region of transition from the stream to the surrounding aquifer, the hyporheic zone has been shown to be a critical component of the environment of stream ecosystems. Sharp gradients in physical, chemical, and biological conditions produce extreme diversity in the hyporheic zone, and often give rise to unique processes that do not occur elsewhere in the overlying stream or underlying aquifer. As a result, fluxes through the hyporheic zone tend to influence a very wide range of ecologically relevant substances, including nutrients, carbon, and contaminants (Brunke & Gonser, 1997; Winter et al., 1998; Jones & Mulholland, 2000). However, there is currently great uncertainty about the relationship between stream conditions, hyporheic exchange fluxes, and overall transport of ecologically relevant substances (Jones & Mulholland, 2000; Hall et al., 2002).

A considerable amount of recent research has sought to elucidate the basic hydrodynamic processes that are responsible for hyporheic exchange. Speaking broadly, this work can be subdivided into field studies to examine the features of the stream-aquifer system that drive exchange, and laboratory studies to examine the underlying processes in detail. Herein, we synthesize much of the recent process-oriented laboratory work so as to draw some general conclusions about the factors that control hyporheic exchange. We will analyze the role of stream conditions in driving exchange fluxes and the role of sedimentary conditions in establishing the pore water transport environment. By considering the functionality of exchange fluxes with stream and sedimentary parameters, we will attempt to identify key variables that are expected to be the most critical determinants of hyporheic exchange. This information should clarify the underlying hydrodynamic and hydrogeologic factors that control hyporheic exchange and prove useful to stream ecologists and other stream researchers in comparing results obtained from different study sites or at one site under different flow conditions, designing experiments to measure hyporheic processes, and distinguishing the rates of hyporheic biological or biogeochemical processes from hydrodynamic exchange rates.

Theory

Sandy stream beds are generally covered by periodic topographical features known as bedforms. These bedforms develop characteristic shapes due to the interplay of the stream flow and transport of bed sediment grains. Protrusion of bedforms into the stream flow causes a periodic variation in the dynamic head at the bed surface, with high pressure occurring on the upstream sides of bedforms and low pressure in the recirculation zone downstream of bedforms. These pressure gradients drive advective pore water flow into, through, and out of the bed. In sandy sediments, exchange due to these advective hyporheic exchange flows generally dominates diffusive transport by two orders of magnitude or more (Savant et al., 1987).

Elliot & Brooks (1997a) analyzed this advective 'pumping' process from fundamental hydrodynamic principles. Their analysis involved approximating the boundary head distribution with a sinusoidal profile, solving Laplace's equation in the subsurface to determine the pore water head distribution, and applying Darcy's law to determine the advective pore water velocity field. The characteristic Darcy velocity for hyporheic exchange is given by $u_m = kKh_m$, where k is the wave number of the bedforms ($k = 2\pi/\lambda$, where λ is the wavelength), K is the hydraulic conductivity of the sediments, and h_m is the half-amplitude of the sinusoidal head distribution at the bed surface. The variation of head over the bedform surface depends on the velocity head of the stream ($V^2/2g$) and the extent of bedform penetration into the stream (H/d) according to:

$$h_m = 0.28\frac{V^2}{2g}\begin{cases}\left(\frac{H/d}{0.34}\right)^{3/8} & \text{for } H/d \leq 0.34\\ \left(\frac{H/d}{0.34}\right)^{3/2} & \text{for } H/d > 0.34\end{cases} \quad (1)$$

where V is the stream velocity, g is the acceleration due to gravity, H is the bedform height, and d is the stream depth.

The pore water velocity field can be used to directly calculate hyporheic exchange. The characteristic dimensionless timescale for bedform-driven advective exchange with an infinite sediment bed is given by Elliott & Brooks (*loc. cit.*):

$$t^* = ku_m\frac{t}{\theta} = k^2Kh_m\frac{t}{\theta} = 0.28\,\frac{k^2KT}{\theta}\frac{V^2}{2g}\left(\frac{H/d}{0.34}\right)^{3/8} \quad \text{for } H/d \leq 0.34 \quad (2)$$

where θ is the porosity of the bed sediments. This theory has been extended to allow analysis of exchange with finite stream beds as well (Packman et al., 2000a). For shallow stream beds where an impermeable layer constrains exchange to a depth, d_b, the appropriate exchange timescale is:

$$t_f{}^* = \frac{t^*\tanh(kd_b)}{kdb} = 0.28\,\frac{kKt}{\theta}\frac{\tanh(kd_b)}{d_b}\frac{V^2}{2g}\left(\frac{H/d}{0.34}\right)^{3/8} \quad \text{for } H/d \leq 0.34 \quad (3)$$

This theory indicates that the rate of hyporheic pumping exchange is strongly dependent on the stream velocity (V), the bedform geometry (particularly the wavelength $\lambda = 2\pi/k$), and the hydraulic conductivity (K) and porosity (θ) of the bed sediments. In addition, the depth of the sediments is also important when hyporheic exchange is constrained by an impermeable boundary.

The exchange in a gravel bed can be significantly enhanced relative to that in a sand bed. In addition to bedform-induced advective flows, direct coupling of stream and pore water flows due to turbulent interactions can produce significant exchange in gravel bed sediments (Nagaoka & Ohgaki, 1990). Zhou & Mendoza (1993) theorized that the larger pore spaces in gravel admit the penetration of considerable stream-driven turbulence into the bed, which causes a non-Darcy slip velocity at the bed surface and an exponentially decreasing flow in the bed. We recently compared hyporheic exchange in gravel beds with and without bedforms (Packman et al., 2003), and found that net exchange would often be the product of both advective and diffusive processes operating

Table 1. Experiments used in our evaluation of hyporheic exchange

Data Source	Run numbers	Sediment properties			Tracer
		d_g (mm)	K (cm/s)	θ (–)	
Elliot & Brooks (1997)	17	0.13	0.0079	0.295	Fluorescein
	8–10, 12, 14–16	0.48	0.11	0.325	
Eylers et al. (1995)	11	0.20	0.04	0.325	Lithium
Packman et al. (2000)	10, 12–15, 17	0.48	0.15	0.325	Lithium
Packman & MacKay (2002)	1–3*	0.48	0.18	0.325	NaCl
Marion et al. (2002)	2–3	0.85	0.38	0.38	NaCl
Packman et al. (2002)	1–11	5.0	15.0	0.38	NaCl
Nagaoka & Ohgaki (1990)	1–3, 5–7, 9, 13	40.8	–	0.362	NaCl

* Each of these experiments involved two measurements of hyporheic exchange (clean and clogged bed conditions).
– Not reported.

simultaneously. Exchange with bedforms appeared to be enhanced due to both turbulent transport in the upper layers of the bed, and advective hyporheic flow patterns were observed even when the bed was flat, apparently due to periodic pressure variations on the bed surface produced by subtle variations in the bed surface on the order of the sediment grain diameter.

Because of the complexity of the hyporheic flows in gravel stream beds, no fundamentally based theory exists to predict the resulting hyporheic exchange in terms of measured stream and bed parameters. Instead, bulk solute exchange can be modeled with an effective diffusion coefficient. For the case where there is initially a tracer concentration C_0 in the stream and no tracer in the subsurface, the exchange flux is given by:

$$J = C_0/\pi(D_{\mathrm{eff}}/t)^{1/2}, \qquad (4)$$

where D_{eff} is the effective diffusion coefficient. Using this relationship, the magnitude of hyporheic exchange flux can be consistently obtained from the initial decrease of the in-stream solute concentration measurements as an effective diffusion coefficient.

It should be emphasized that application of this diffusive relationship does not imply that the exchange necessarily occurs due to diffusive transport processes. An effective diffusion coefficient is utilized here as a convenient idealized representation of hyporheic exchange that allows comparison of the relative rates of exchange that occur under different conditions and due to different mechanisms. We expect that the stream flow conditions (velocity and depth), sedimentary conditions (permeability and porosity) and bed topography (such as the bedform height and wavelength)

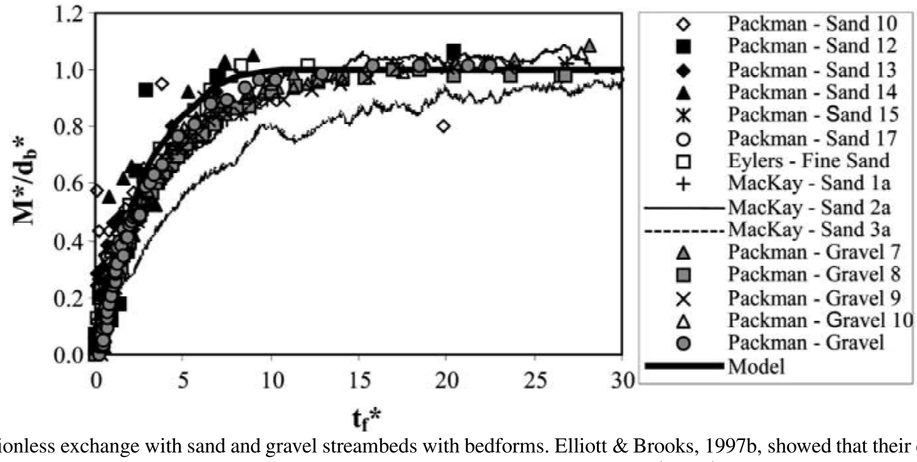
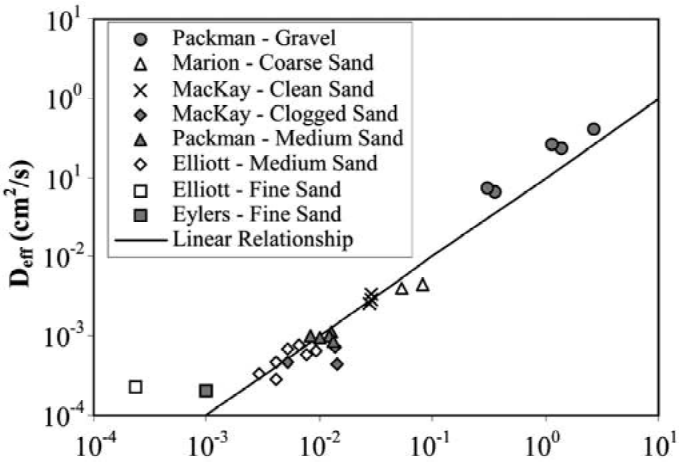


Figure 1. Dimensionless exchange with sand and gravel streambeds with bedforms. Elliott & Brooks, 1997b, showed that their data on exchange with deep stream beds would collapse similarly when plotted as a dimensionless penetration M^* vs. t^*.



$$\frac{Kh_m}{\theta} = 0.28\frac{K}{\theta}\frac{V^2}{2g}\left(\frac{H/d}{0.34}\right)^r \qquad r = 3/8 \text{ or } 3/2$$

Figure 2. Effective diffusion coefficients for experiments with bedforms compared with pumping theory. Bedform-induced hyporheic exchange is shown to be influenced by the stream velocity, the permeability and porosity of the bed sediments, and the relative bedform height.

will generally control hyporheic exchange. By applying the preceding analysis to hyporheic exchange observed under a variety of conditions, we will clarify the relative roles of stream and sediment conditions in controlling hyporheic exchange.

Materials and methods

Exchange between streams and streambeds has been studied in several different recirculating flumes using similar methods (Nagaoka & Ohgaki, 1990; Eylers et al., 1995; Elliot & Brooks, 1997b; Packman et al., 2000b; Packman & MacKay, 2003; Marion et al., 2002; Packman et al., 2003). Flume experiments offer many advantages for the study of hyporheic exchange. Primarily, they allow examination of exchange processes under controlled, easily observable stream and subsurface conditions. Full control of the stream and sedimentary conditions allows examination of the importance of individual processes or experimental variables. By synthesizing results from a number of different studies, we seek herein to parameterize general relationships for stream-side and sediment-side control of hyporheic exchange. The experiments that will be used for this comparison, summarized in Table 1, used bed sediments ranging from fine sand (d_g = 130 μm) to coarse gravel (d_g = 4.08 cm), stream velocities between 4 and 40 cm/s, and different bed geometries including both flat beds and a variety of sizes of bedforms. Detailed information on the conditions of individual experiments can be found in the primary citations indicated in Table 1.

Similar experimental methods were used for the first six studies listed in Table 1: (i) the desired bed topography was produced either artificially or naturally at an appropriate stream velocity; (ii) steady, uniform flow was established over a stable sediment bed (with or without bedforms), (iii) a conservative tracer was added uniformly to the recirculating stream in the flume, (iv) the tracer concentration in the stream water was measured over time. Sodium chloride (NaCl) was the most frequently used conservative tracer, but lithium ion and a fluorescent dye were also used. For this methodology, the net tracer exchange with the bed is calculated from the reduction in the in-stream concentration over time by considering the mass balance between the recirculating stream and pore water in the sediment bed.

The study of Packman & MacKay (2003) employed this methodology, but also involved observations of the deposition of considerable quantities of kaolinite clay in the streambed. Conservative solute injections were used to assess hyporheic exchange both before and after the streambed became clogged with clay. Each of these tracer injections will be considered separately here.

Nagaoka & Ohgaki (1990) used an entirely different experimental procedure: they injected a sodium chloride solution directly into the stream bed and ob-

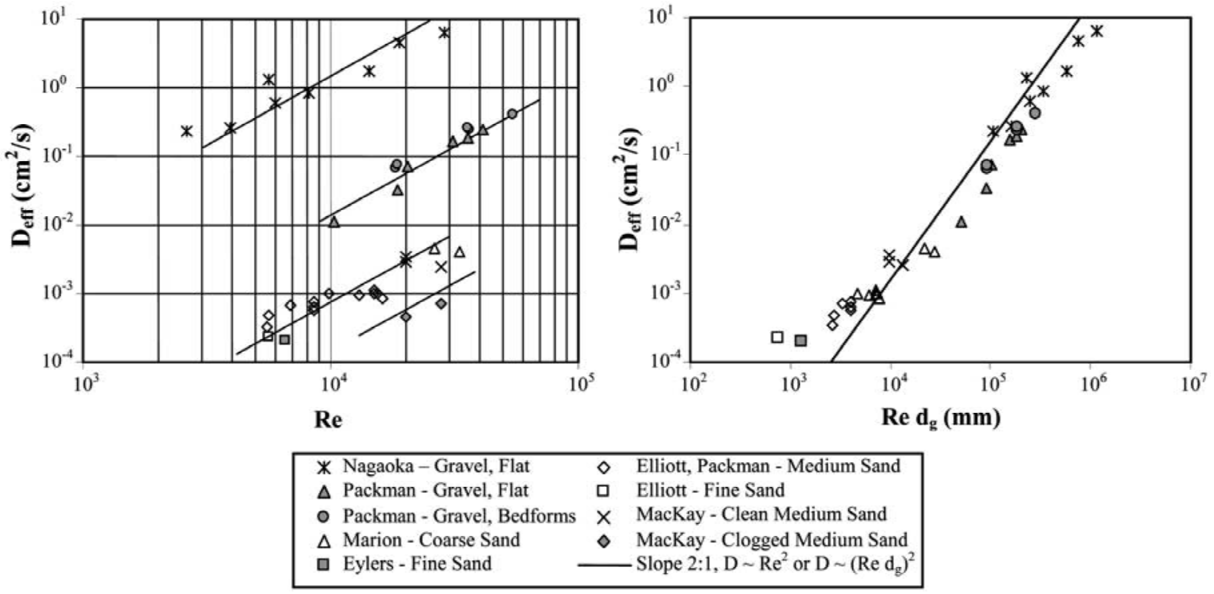


Figure 3. Comparison of effective diffusion coefficients for hyporheic exchange with both flat beds and bedforms. Left panel shows that the hyporheic exchange rate is proportional to the square of the stream Reynolds Number. Right panel shows that exchange is also proportional to the square of the sediment size, which suggests that the permeability of the sediment bed controls exchange with both flat beds and bedforms.

served the resulting tracer concentration at several downstream locations in the bed. This procedure did not allow observation of net hyporheic exchange, but did yield measurements of the effective diffusion coefficient for exchange across the stream-subsurface interface. These diffusion coefficients will be compared with the effective diffusion coefficients obtained from the other experiments.

Results

Exchange results will be compared in two different ways. Firstly, we will examine the ability of advective pumping theory to predict the exchange induced by bedforms. In this comparison, we will use the in-stream tracer concentration data measured in each experiment. This comparison will be restricted to cases with bedforms, as no equivalent theory exists to attempt to predict exchange with flat sediment beds. Secondly, we will compare the results of all experiments by using the effective coefficient to represent the hyporheic exchange rate in each experiment. We will particularly identify relationships between observed exchange rates and basic hydrodynamic and sedimentary parameters.

In comparing experimental results obtained with different system geometries, it is convenient to represent hyporheic exchange as the average depth of solute penetration into the subsurface, M. As the experiment proceeds, tracer from the stream increasingly mixes into pore water in the stream bed, so that the average depth of solute penetration is a surrogate parameter for the total mass of tracer that has been transferred to the bed. Exchange results from a number of different experiments with bedforms are presented in Figure 1 as the fraction of the bed exchanged, M/d_b, plotted against the appropriate dimensionless time scale for exchange with shallow beds. It can be seen that the dimensionless theoretical model prediction does a good job of representing almost all of the experimental results. It should be emphasized that the model was not fit to any of the results for exchange with clean sediment beds. The results of Packman & MacKay's (2003) experiments on exchange with clogged streambeds were modeled by adjusting the hydraulic conductivity and porosity, since these conditions were not known after clay plugged the bed.

The fact that a single, dimensionless theoretical solution can predict such a wide variety of observed exchange behavior shows that the underlying mechanism of advective hyporheic exchange is well-understood, and that the quantitative description for hyporheic flows induced by common dune-shaped bedforms applies for a wide range of stream and sedimentary conditions. The same model cannot necessarily be expected to apply directly to natural streams, which can have a range of different bed features. Even so, the underlying process of hydrodynamic interactions between the stream- and pore-water flows will drive exchange with other stream channel features as well, which suggests that elements of this theory should be able to be used to interpret hyporheic exchange for a very wide range of natural conditions.

The general applicability of pumping theory can be assessed by evaluating how the exchange flux scales with important stream parameters. In Figure 2, we show that there is a simple linear relationship between the effective diffusion coefficient for exchange and the parameter grouping Kh_m/θ. Note that the theoretical curve indicates a linear relationship between D_{eff} and Kh_m/θ, but we plot this on a log-log scale in order to show a wide range of experimental data in one plot. The compilation of results shown in Figure 2 indicates that this linear relationship holds for more than three orders of magnitude of bedform-driven exchange. The parameters K and θ are the sediment permeability and porosity, respectively, which characterize sedimentary hydrogeologic control of hyporheic exchange. The parameter h_m reflects the bedform-induced pressure variation at the bed surface that drives advective exchange flows. As indicated by Equation (1), which applies for dunes, the driving mechanism for exchange includes both the velocity head of the stream ($V^2/2g$) and a geometric term related to the dune shape. While the exact relationship given in Equation (1) only holds for a specific bed geometry, albeit a common one, the dependence on V^2 and K/θ reflects underlying hydrodynamic processes that control pore water flow.

To show the generality of this approach, in Figure 3 we examine the dependence of exchange with stream and sedimentary properties for the full set of experimental data. Even though different exchange mechanisms are reflected in the experiments conducted with flat beds and bedforms, and with sand and gravel sediments, the left panel of Figure 3 demonstrates that all observed effective diffusion coefficients scale with the square of the stream Reynolds Number (Re). Since Re = Vd/η, where d is the stream depth and η the kinematic viscosity of water, these results indicate that exchange is proportional to V^2 for all conditions tested. Individual data sets have different intercepts in the left panel of Figure 3 due to the effect of the sediment properties on the exchange rate.

The right panel of Figure 3 shows that the exchange rate also scales with the square of the sediment grain diameter. This indicates that the sediment grain diameter can be effectively used as a surrogate for the hydrogeologic properties that depend on it, namely K and θ. The observed dependence on $d_g{}^2$ was expected because the intrinsic permeability of a granular porous medium is known to be approximately proportional to the square of the grain diameter. Nonetheless, we find it remarkable that the linear relationship between exchange rate and the parameter grouping (Re d_g)^2 holds so well for almost five orders of magnitude of observed hyporheic exchange behavior, representing two orders of magnitude of variation in the sediment grain size, almost an order of magnitude variation in stream velocity, and distinctly different stream channel topographies. The success of the scalings presented in Figures 2 and 3 suggests that these relationships can be applied to a wide range of natural conditions.

Discussion and conclusions

In this work, we synthesized and compared the results of a wide variety of laboratory experiments that had been used to probe hyporheic exchange processes. All of these experiments measured the rate of exchange across the stream-subsurface interface, and most of them also evaluated the bulk transfer of a conservative solute from the stream to the hyporheic zone. Experimental conditions covered a range of sediment sizes from fine sand (d_g = 130 μm) to gravel (d_g = 4.08 cm), stream velocities from 4 to 40 cm/s, stream depths from 3 to 20 cm, and bed topography ranging from flat beds to 32 cm long and 4 cm tall dunes. Experimental results were compared both in terms of the complete exchange curves (dimensionless mass transfer vs. dimensionless time) and by using an effective diffusion coefficient to represent the exchange rates of experiments conducted under disparate conditions.

This comparison clearly showed the relative effects of stream and sedimentary conditions in controlling hyporheic exchange. Exchange due to bedforms is predicted well by the fundamentally derived advective pumping model. Interactions between the stream flow, bedform topography, and pore water causes the hyporheic exchange rate to be proportional to the permeability of the bed sediments and the square of the stream velocity, and inversely proportional to the depth of the bed and the porosity of the sediments. In addition, the exchange rate also depends on the bedform geometry, and particularly on the relative penetration of the bedform into the stream.

While predictive theoretical analysis is currently limited to specific classes of bedform geometry, the fundamental nature of the stream-subsurface flow interactions indicates that the hyporheic exchange rate should generally be proportional to the permeability of the sediments and the square of the stream velocity. All experimental exchange results evaluated here were shown to scale with these parameters. These two parameters can thus be considered the most critical in attempting to assess the hyporheic exchange rate in any natural stream. The stream velocity is important because coupling between the stream flow and sedimentary pore water will often drive hyporheic exchange, and the hydraulic conductivity controls the ability of the sediments to admit any sort of advective flux.

These results can be used to improve interpretation of field measurements of hyporheic exchange. This work indicates that scaling arguments can be used to compare hyporheic exchange rates measured at different sites or to apply measurements made at a site to different flow conditions. The primary advantage of this is that the effects of simple flow variations can be isolated so that differences in the underlying hyporheic conditions can be assessed. We previously presented a methodology for this in Wörman et al. (2002) and showed in Salehin et al. (2003) that this methodology can be used to differentiate variations in the hyporheic zone related to land use and channel characteristics. While those efforts were restricted to a particular stream, the results presented here indicate that similar methods can be successfully applied to a very wide range of stream and sedimentary conditions.

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